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# SATELLITE STUDIES OF CLOUDS AND CLOUD BANDS NEAR THE LOW-LEVEL JET [1]

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ABSTRACT

TIROS–VII and –VIII photographs are used to study synoptic and mesoscale cloud patterns during 6 days with strong southerly flow in the Central United States. Cases with nocturnal thunderstorms show a tendency for thunderstorms to occur in the downstream portion of the jet. Through advection of moisture and turbulent breakdown of the nocturnal inversion, the low-level jet plays an important role both in the formation of stratus and in its manner of dissipation. Longitudinal cumulus cloud bands with spacing of 10 to 15 km occur with slightly superadiabatic lapse rates and moderate wind shear in agreement with theoretical results by Kuo.

## 1. INTRODUCTION

Previous studies (Newton, 1956; Wexler, 1961; Bonner, 1965*a*) have shown that the low-level jet stream in the Central United States is, at times, a widespread and well-organized phenomenon.

Typically, the jet occurs with strong, southerly geostrophic winds. Its altitude varies from about 300 m to 1500 m above the ground. Wind speed in the jet frequently reaches 25 to 30 m sec$^{-1}$, and wind shears within the boundary layer may be as large as 20 to 40 m sec$^{-1}$ per km (Bonner, 1965*a*). Although distinct low-level wind maxima may be found on both daytime and nighttime soundings, the jet is generally strongest and of greatest horizontal extent from about midnight to 06 local time.

The major purpose of this study is to examine, from satellite photographs, the cloud patterns in the vicinity of the low-level jet. The jet has long been considered an important factor in thunderstorm formation in the Midwest (Means, 1952)—and, in particular, in the formation of nocturnal thunderstorms in the region around Iowa and Nebraska (Pitchford and London, 1962). In addition, since wind shear is considered to be a primary factor in the formation of cumulus cloud bands (see, for example, Malkus, 1963), it seemed logical that an examination of cases of strong southerly flow over the Central United States might provide an excellent opportunity for examining the conditions for band formation. Satellite photographs show that small-scale bands—wavelengths of the order of 10 km—are much more common over oceans than over land. However, studies of bands over the oceans (Schuetz and Fritz, 1961; Malkus, 1963) are handicapped by the fact that very little surface data are available to aid in the identification of cloud types and altitudes, and virtually no information is available about such important parameters as wind shear and thermal stability. Observations from island stations, while useful, may not represent the actual structure of the boundary layer over the ocean.

Thus, we decided to examine first the broad-scale cloud patterns in the vicinity of the low-level jet—relating these to moisture, stability, and large-scale vertical motions—and second, to look for cases of banded convection.

## 2. APPROACH

Six cases were selected from a sample of spring and summer days in 1964 for which TIROS–VII or –VIII photographs were available. In all cases there was strong southerly or southwesterly flow over the Central United States. Picture times on the various orbits ranged from about 07 CST to 16 CST. Although it would have been interesting to examine the cloud patterns at night when the jet is best developed, the medium resolution infrared data from TIROS in the standard format are not of sufficiently high resolution to identify small-scale cumulus bands. No usable situations were found within the limited period for which Nimbus–I high resolution infrared data are available.

Three morning and three afternoon cases were studied. On two of the morning cases there were strong southerly jets with wind speeds of 50 to 60 kt within the lowest kilometer. Table 1 lists the cases and corresponding satellite orbits. Wind and temperature soundings were plotted at all stations in the region of interest. Radiosonde data, hourly airways reports, and facsimile radar summaries were used in the cloud analysis.

TABLE 1.—*Cases and satellite orbits*

| Date | TIROS | Orbit | Time (GMT) |
|---|---|---|---|
| Feb. 29, 1964 | VIII | 1023/22 | 2200 |
| Apr. 27, 1964 | VII | 4632/31 | 1637 |
| May 4, 1964 | VII | 4734/33 | 1414 |
| May 5, 1964 | VII | 4749/48 | 1435 |
| | VII | 4748/47 | 1259 |
| May 7, 1964 | VII | 4788/77 | 1345 |
| | VIII | 2006/05 | 1704 |
| May 26, 1964 | VIII | 2283/82 | 2010 |

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### PICTURE GRIDDING

Because of our interest in the mesoscale and in using surface reports to aid in interpreting the pictures, it was important to grid the photographs carefully. Grids were computed with a machine version of the Fujita technique (Bonner, 1965*b*). Input consists of the subpoint track, satellite altitude, picture start time, and the location of the spin-axis point at some reference time. For several orbits, picture start times and spin-axis points were carefully determined from the photographs in the manner described by Fujita (1963); in others, we used published spin-axis data and made corrections where necessary to match any landmarks appearing on the pictures. Program output is a matrix of coordinates of 2° latitude and longitude intersections on a distortion-free grid that must then be transformed by hand to the picture itself. The method is laborious, but it is less so than the purely graphical technique and is fully as accurate.

### VERTICAL VELOCITIES

Vertical velocities were computed from a three-level, quasi-geostrophic model using input levels of 850, 500, and 200 mb. Equations and boundary conditions are described in the Appendix, and it should be sufficient to mention here that the lower boundary condition includes both the effects of terrain slope and frictional convergence. We experimented with a four-level model having a kinematically determined lower boundary condition at 800 mb. While this seemed to give a more realistic representation of vertical velocities along the mountain slopes, there were problems with the model; and the three-level formulation was used in all case studies presented here. Output is in microbars sec$^{-1}$ at 675 and 350 mb.

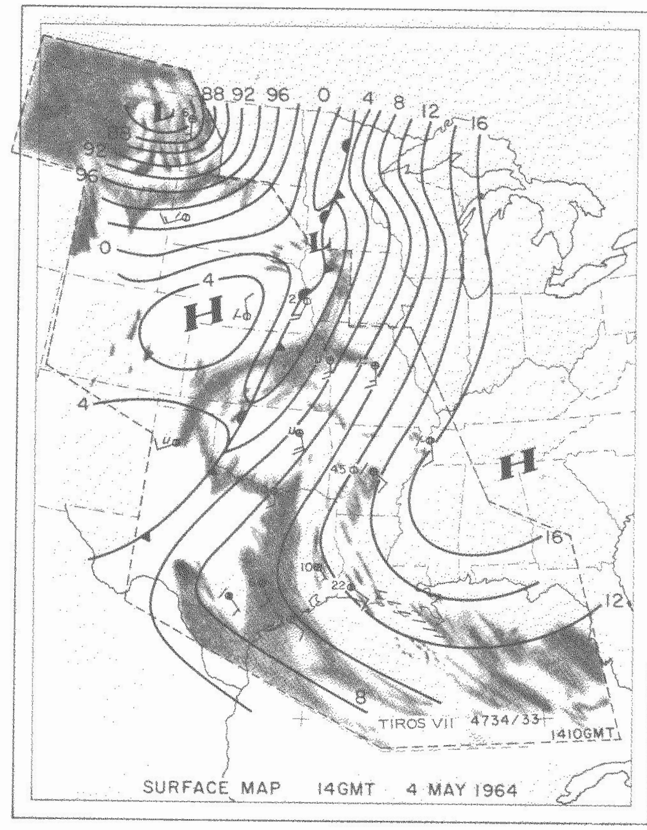


FIGURE 1.—Surface map at 14 GMT on May 4, 1964. Representative surface observations of wind, cloud amount, and cloud-base height are plotted in standard airways format. For example, 30 followed by a circle circumscribing a plus sign means overcast sky with cloud bases at 3,000 ft. Shaded areas indicate regions of cloud cover as determined from satellite photographs.

## 3. BROAD-SCALE CLOUD PATTERNS

### CASE 1: MAY 4, 1964

Figure 1 shows the surface map at 14 GMT—roughly 15 min before the time of the satellite pictures. In all of the cases to be shown, hand-drawn cloud mosaics were constructed as overlays to the weather maps. Shaded regions of the map correspond to cloud areas on the individual photographs.

There are two main regions of cloudiness. One is in Montana—over and to the west of a deep, occluded cyclone. The second is in Texas. Clouds in Montana are relatively bright, and the surface map indicates ceilings in this area generally below 1,000 ft with steady rain at most stations. The clouds in Texas are stratus with tops between 3,000 and 5,000 ft.

The low-level jet is well developed at 12 GMT with southerly winds of 50 kt or more from east-central Texas to eastern Kansas (fig. 2). At the core of the jet, the only clouds are cirrus. There are, however, two features of the large-scale cloud distributions that do appear to be related to the jet:

1) The extension of the stratus in a narrow arm through Texas—almost exactly along the axis of the jet.

2) The occurrence of scattered early morning thunderstorms downstream from the jet core in southern Minnesota and extreme northern Iowa. This is just beyond the area of picture coverage—except for a region of relatively bright clouds in northern Iowa—but the thunderstorm activity is documented by radar summaries.

Figure 2 shows the vertical motions at 675 mb and mixing ratios at 850 mb—with cloud patterns superimposed. Broad-scale clear and cloudy areas along the western part of the map correspond quite closely to the centers of rising and sinking motion. Thunderstorms in Minnesota occur within a weak center of low-level convergence—$0.5\times10^{-5}$ sec$^{-1}$ as determined from the three-level model and approximately $1\times10^{-5}$ sec$^{-1}$ from the four-level vertical velocities.

The moist tongue in figure 2 is nearly parallel to and slightly to the right of the jet at the 700-m level.

### CASE 2: MAY 5, 1964

TIROS VII passed over the area on two successive orbits—at 1259 GMT and 1435 GMT. The surface map shown in figure 3 is for 15 GMT; however, plotted data correspond to the hour closest to the picture times.

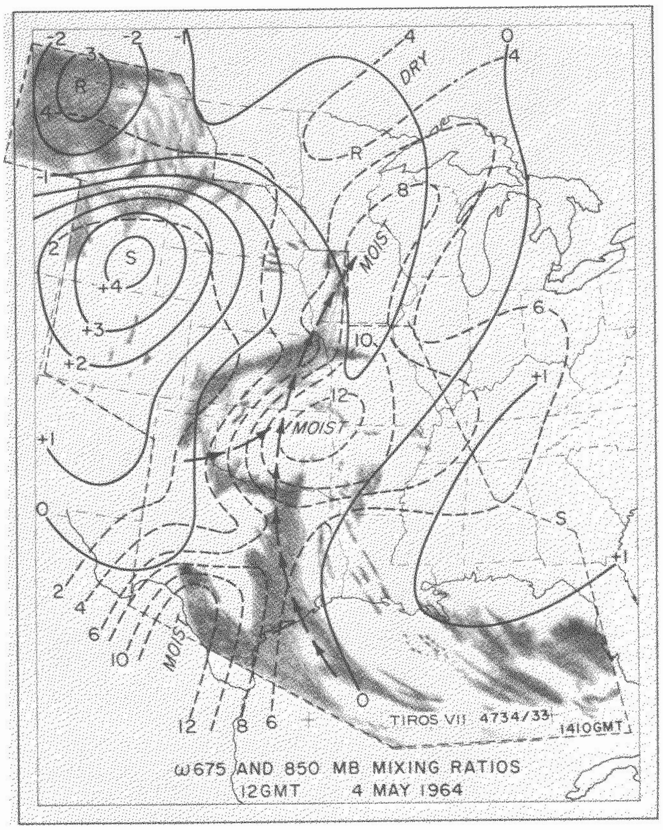


FIGURE 2.—Vertical motions at 675 mb (solid lines) and 850-mb mixing ratios (dashed lines); ω is in microbars per second; mixing ratios in grams per kilogram. Heavy arrows show position and direction of the jet at 700 m above the ground.

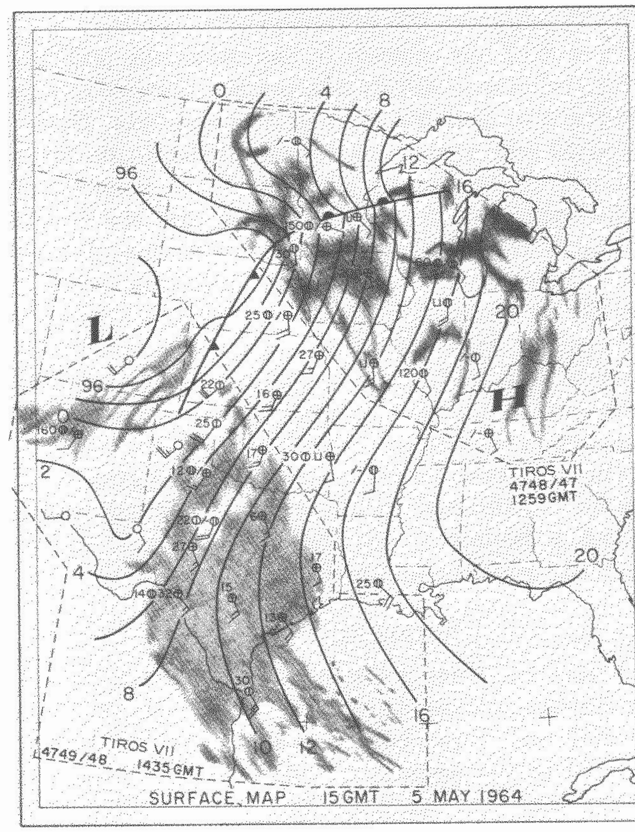


FIGURE 3.—Surface map at 15 GMT on May 5, 1964. Cloud cover from TIROS orbits at 1259 and 1435 GMT. Explanation as in figure 1.

The low-pressure center over Montana on May 4 has moved northward into Canada. The zone of sinking motion south of the Low now appears only as a weak center on the northern boundary of the map. The front has shifted slightly to the west in Kansas and Nebraska, and a weak cyclone is developing in Colorado.

There is an area of middle and high cloudiness in the upper Midwest. Once again, an extensive stratus overcast covers much of Texas reaching into Oklahoma and Kansas and, as stratocumulus, into southwestern Nebraska.

Vertical velocities (fig. 4) are generally less than 1 cm sec$^{-1}$. The air is ascending weakly along the axis of the jet (fig. 4) and rising at a rate slightly greater than 1 cm sec$^{-1}$ in the downwind section of the jet. The moist tongue at 850 mb lies approximately along the axis of the jet.

Low-level wind speeds and wind shears are very pronounced along a line from central Texas through Kansas into eastern Nebraska. At Topeka, Kans., the wind is 5 m sec$^{-1}$ at the surface and 28 m sec$^{-1}$ at 700 m above the ground. The stratus again extends northward along the zone of high moisture content, strong winds, and boundary-layer wind shears.

## CASE 3: MAY 7, 1964

Most complete coverage of the area is provided by TIROS VIII, Orbit 2006. The surface map at 17 GMT (fig. 5) corresponds to within 6 min with the times of the pictures used for the cloud analysis.

The front along the western boundary of the region is primarily a zone of strong horizontal gradient of moisture at low levels. The only cloud cover of interest is the stratus and stratocumulus deck extending from Texas to Iowa.

Vertical velocities are again small in the absence of major disturbances. The zero line in figure 6 corresponds closely to the edge of the low and middle clouds in Kansas and Nebraska. Clear skies are found in the zone of weak descent in the upper Midwest. The air is rising near the jet with the maximum rate of ascent occurring in a zone where two jet maxima converge. The westernmost jet is the stronger of the two with winds exceeding 20 m sec$^{-1}$ at 700 m above the ground in a narrow band from northern Texas to eastern Nebraska.

In Texas, the moist tongue lies between the two jet maxima. Moisture patterns at 850 mb follow closely the vertical motion patterns, and it is likely that the extension of the moist tongue into Arkansas and southern Missouri results from an increase in the depth of the moist layer by low-level convergence. The strong gradient of mixing ratios in Kansas and Missouri results from a lowering of the inversion to below 850 mb and does not correspond to a boundary in the clouds. Broad-scale cloud features are similar to those on May 5 (fig. 3), but the stratus now extends farther to the east in Oklahoma and Texas, along the easternmost of the two low-level jets.

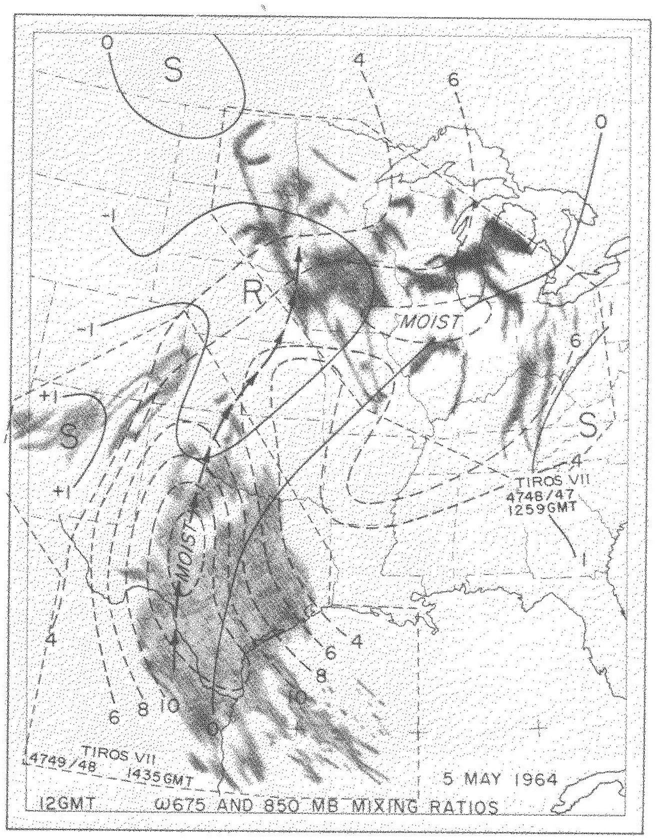


FIGURE 4.—Vertical motions and mixing ratios. Explanation as in figure 2.

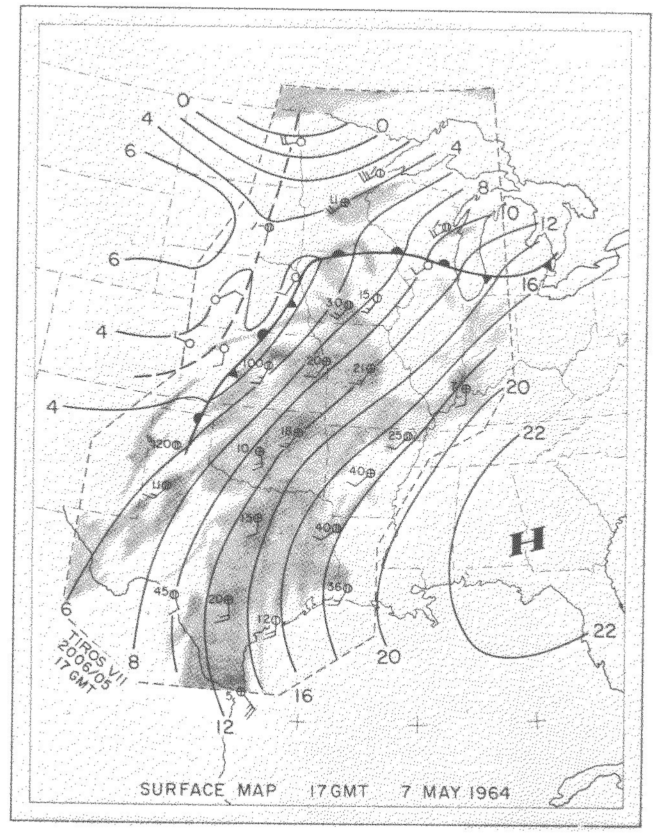


FIGURE 5.—Surface map at 17 GMT on May 7, 1964. Explanation as in figure 1.

## CASE 4: MAY 26, 1964

Picture times and map time correspond in this case to within about 12 min. Figure 7 shows a weak cold front at 20 GMT extending through northern Missouri and central Kansas. The low-pressure center in Kansas is not accompanied by a closed circulation and represents at this time simply a zone of strong low-level convergence.

Radar reports indicate widespread convective activity with scattered showers embedded in middle clouds through central Nebraska, scattered thunderstorms through Illinois and Indiana and in Louisiana and southern Texas. A severe thunderstorm with radar tops to 60,000 ft is reported directly over the front in eastern Kansas.

The 850-mb moisture pattern (fig. 8) shows a moist tongue extending through eastern Texas into Missouri, Illinois, and Indiana. The low-level jet in this afternoon case is not well developed. It lies within the zone of relatively strong pressure gradient in western Texas, terminating near the front in eastern Kansas. The zone of high mixing ratios parallels the jet in Texas but then turns to the northeast through Indiana and Ohio as westerly winds and weak convergence in the warm air mass spread the moisture south of the front.

Figure 9 is a single TIROS frame taken 4 min after the time of the surface map showing a line of individual bright cloud masses along the front in Kansas. The large bright area in northeastern Kansas is the thunderstorm reported by radar. The cloud band is much narrower

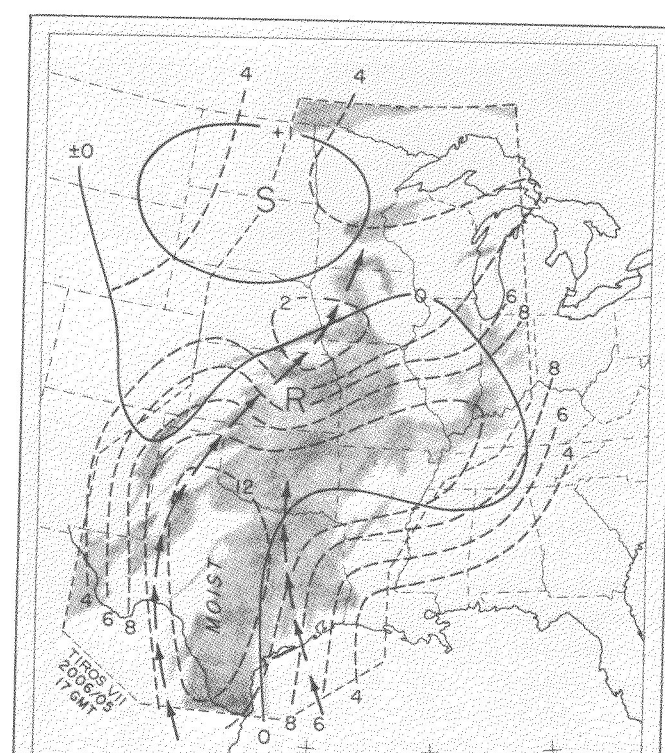


FIGURE 6.—Vertical motions and mixing ratios. Explanation as in figure 2.

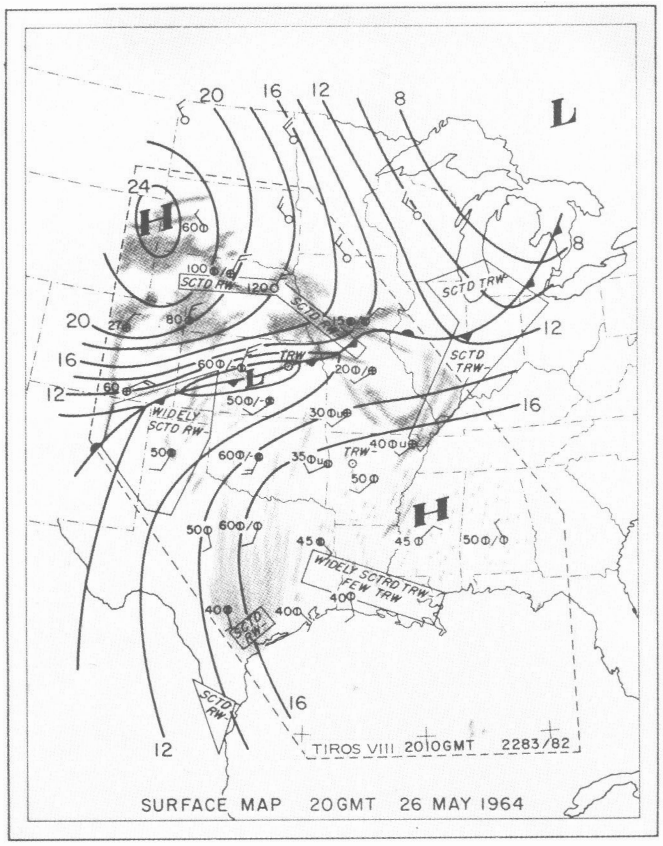


FIGURE 7.—Surface map at 20 GMT on May 26, 1964. Explanation as in figure 1.

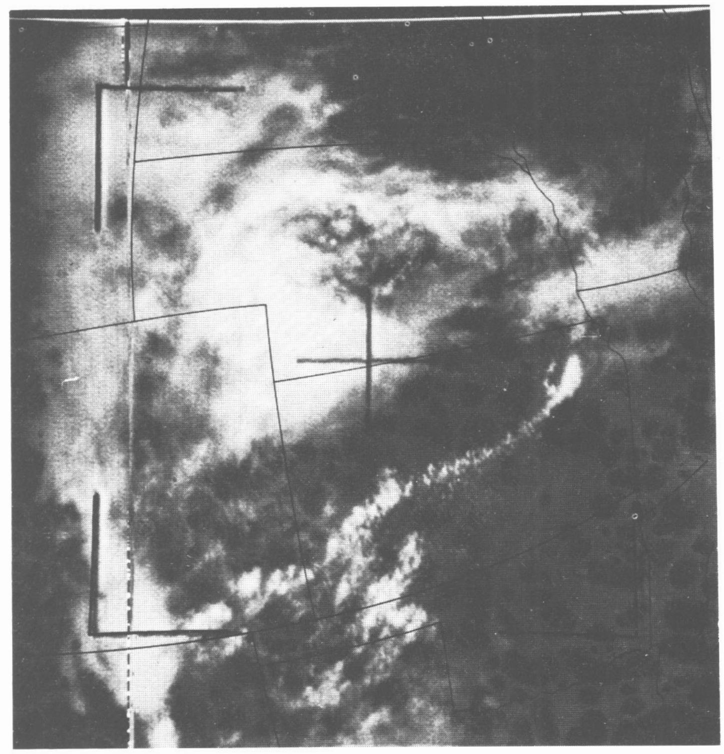

FIGURE 9.—TIROS view at 2004 GMT on May 26, 1964. Note line of cloud elements through Kansas and bright area in northeastern Kansas.

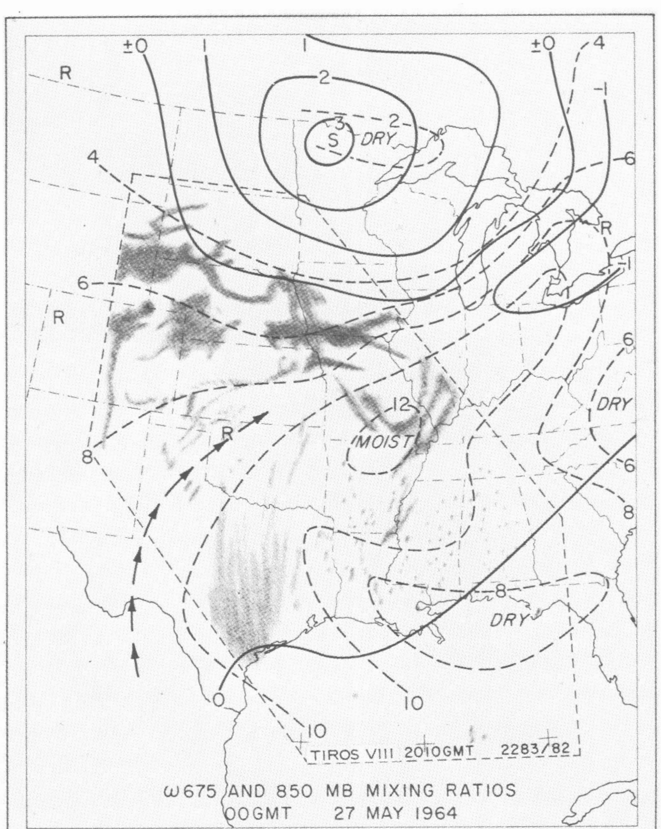


FIGURE 8.—Vertical motions and mixing ratios. Explanation as in figure 2.

than the typical frontal band, and it apparently represents a narrow zone of convergence producing cumulus congestus and, at one point, an intense thunderstorm. It may be significant that this point lies near the intersection of the southerly jet with the front—where the velocity convergence should be a maximum. Vertical velocities in figure 8 do not adequately reflect the convergence in this area because of their scale and because of the highly ageostrophic nature of the flow near the low center.

Cumulus cloud bands in central Texas are oriented parallel to the low-level wind and wind shear.

## CASE 5: APRIL 27, 1964

Figure 10 shows the surface map at 17 GMT—about 23 min after the time of satellite passage.

Strong southerly flow in this case occurs in the region ahead of the front from Alabama to Illinois.

The cloud analysis shows a broad-scale region of cloudiness and rain or snow to the west and northwest of a deep Low in Nebraska. The South Central States are practically clear. Moisture is not being advected inland in Texas by southerly flow, and stratus clouds are found only off the coast. Thunderstorms are reported within a band of clouds ahead of the front in Florida.

The strongest low-level winds in this situation are found to the west and south of the surface Low in a narrow, intense jet of 50-kt winds that follows almost exactly along the edge of the cloud mass in Kansas (fig. 11).

The absence of low clouds over Oklahoma and Texas is easily explained by the large sinking motions and corres-

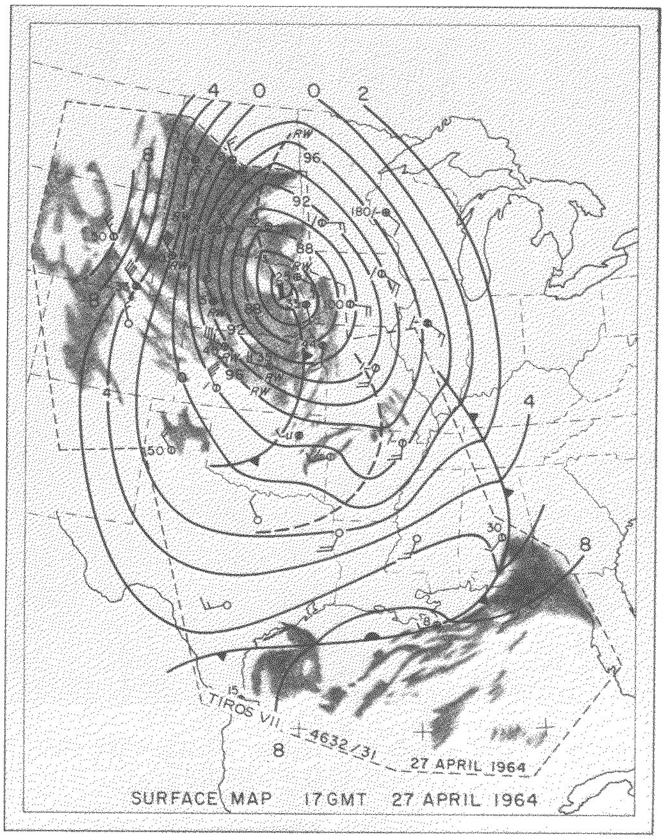


FIGURE 10.—Surface map at 17 GMT on Apr. 27, 1964. Explanation as in figure 1.

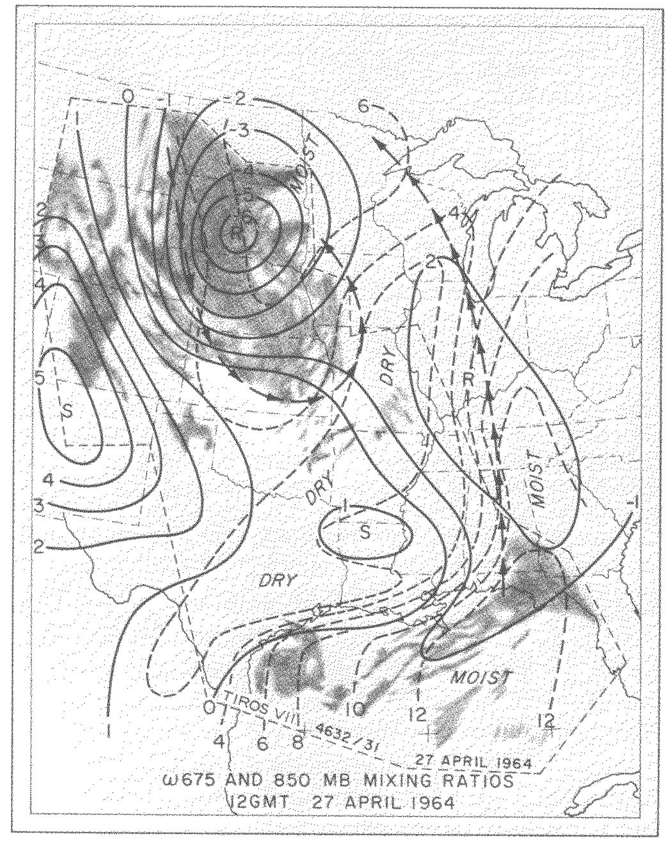


FIGURE 11.—Vertical motions and mixing ratios. Explanation as in figure 2.

ponding dry air in this region (fig. 11). The area of strongest rising motion at 675 mb corresponds closely with the position of the cloud mass and the rain to the west of the surface Low.

### CASE 6: FEBRUARY 29, 1964

The final case to be presented is the only wintertime situation selected. Convective cloudiness associated with the jet is much less frequent in winter than in spring or summer. The case was selected only because of the extremely strong pressure gradient across the Central States.

Time of satellite passage and map time are both 22 GMT (16 CST). Isotachs of 40- and 50-kt winds at 700 m above the ground are indicated on the surface map (fig. 12) to show both the strength and position of the jet in this late afternoon situation. Cloud patterns, mixing ratios, and vertical velocities are not shown since the few significant features can be verbally described.

Scattered-to-broken cumulus and altocumulus extend northward along a moist tongue from Texas into central Kansas. There is no precipitation in the area at the time of satellite passage; however, approximately 8 hr later showers developed in Illinois.

Bonner (1966) and Pitchford and London (1962) show that nocturnal showers and thunderstorms tend to develop in downstream portions of the jet where velocity convergence can lead to synoptic scale ascent of 1 to 10 cm sec$^{-1}$ at approximately 700 mb. Comparison of the position

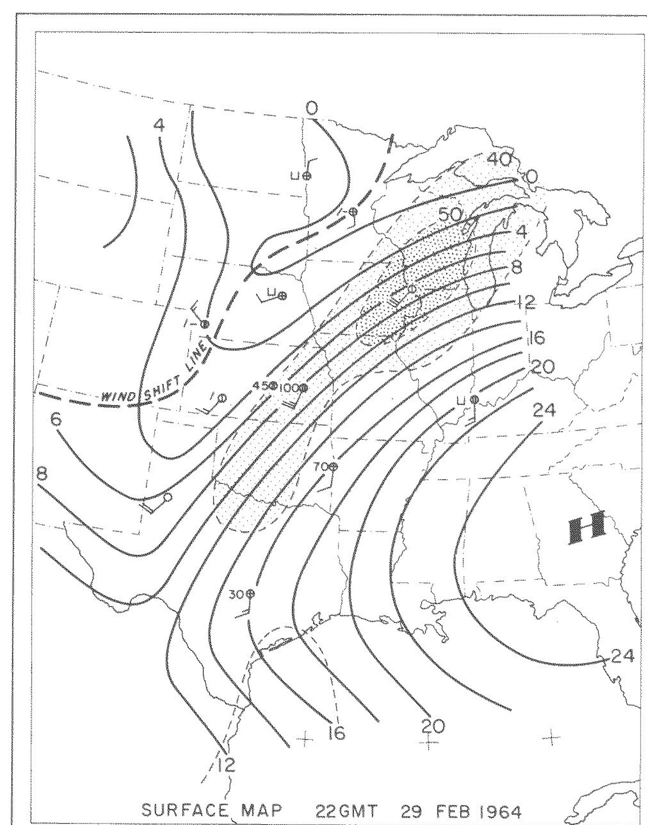


FIGURE 12.—Surface map at 22 GMT on Feb. 29, 1964. Position of the jet is indicated by the shading within 40- and 50-kt isotachs at 700 m.

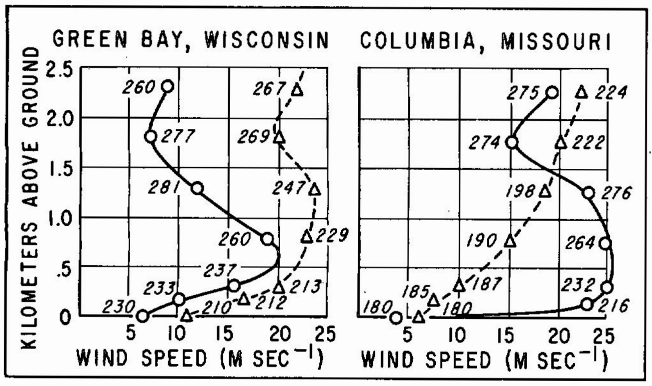


FIGURE 13.—Wind soundings at Green Bay, Wis., and Columbia, Mo., at 00 GMT (dashed lines) and 12 GMT (solid lines) on Mar. 1, 1964. Numbers are wind directions in degrees. Note strong nocturnal increase in speed in lowest 1500 m at Columbia.

of the shower area at midnight with the jet position at 18 CST (fig. 12) indicates that showers develop to the right of the axis, slightly *upstream* from the wind maximum. However, as first shown by Wagner (1939), the nocturnal increase in boundary layer wind speed above the first few tens of meters is normally a maximum in the area from north Texas through Nebraska. Thus, the jet maximum should shift toward the southwest at night. This is exactly what happens, as shown by wind soundings at 18 CST on February 29 and 06 CST on March 1 (fig. 13). As a result, the area of nocturnal showers in Illinois is in the *downstream* or converging section of the jet at midnight.

### SUMMARY

Although six low-level jet cases, with three of them in the same general weather situation (May 4 to 7), cannot define any sort of cloud climatology with respect to the jet, certain common features do show up in the case studies.

First, there has been considerable interest in determining relationships between cloud patterns, as viewed from satellite pictures, and wind, moisture, and vertical motion fields. In those cases presented here, where large-scale disturbances existed (May 4 and April 27), vertical velocities at 675 mb exceeded several centimeters per second in certain areas, and these areas corresponded closely to the areas of broad-scale cloudiness and clear weather. Exact correspondence should not be expected because of errors inherent in the vertical velocity calculations and, more important, because of the physical reality that clouds are transported horizontally from their regions of formation (see, for example, Sanders, 1963). Also, in order to infer anything at all about vertical motions in the middle troposphere or integrated moisture patterns in data-free areas, it is absolutely essential to distinguish between low inversion-type clouds and clouds associated with synoptic-scale disturbances—a distinction that is frequently difficult to make from the pictures alone.

Second, nocturnal showers or thunderstorms were restricted to the downstreams sections of the jet (May 4,

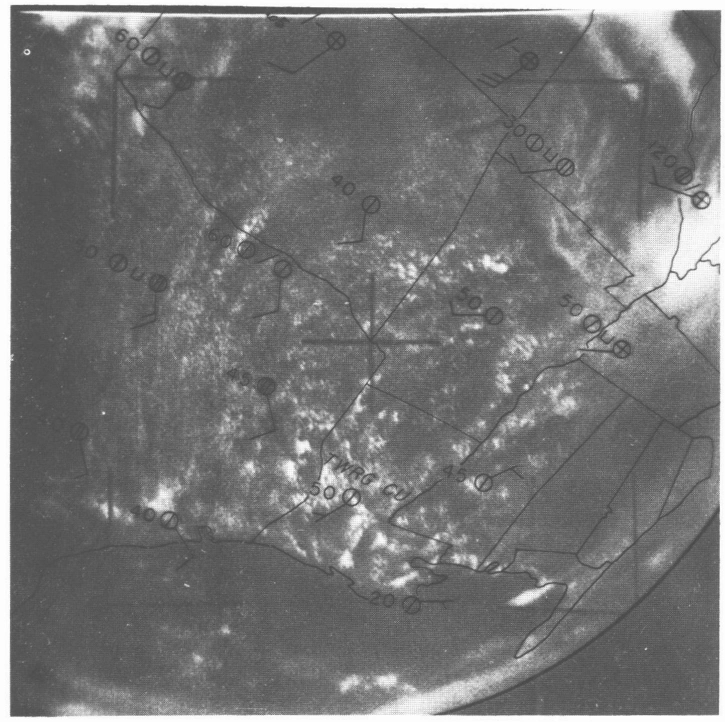

FIGURE 14.—Cloud streets over Texas. TIROS-VIII view, 2006 GMT on May 26, 1964.

February 29), supporting previous conclusions about their mechanism of formation.

Third, stratus decks tended to extend far inland along the axis of strong low-level winds, suggesting a relationship between the low-level jet and the intrusion of Gulf stratus into northern Texas and Oklahoma. The formation of stratus in these regions is probably not the result of simple advection of low clouds but, instead, is produced by turbulent breakdown of the nocturnal inversion under the influence of strong boundary-layer wind shear (Gifford, 1952).

## 4. MESOSCALE PATTERNS

### CUMULUS BANDS

One of the aims of the study was to look for small-scale cumulus bands which we thought should arise in the region of strong wind shear and pronounced curvature of the wind profiles in the low-level jet. Such bands did not appear in the TIROS-VII and -VIII photographs. Instead, the only obvious cases of banded convection occurred in regions of weak to moderate wind shear—away from the core of the jet.

On May 26 (fig. 14, see also fig. 7) there is a faint, streaked pattern of cloudiness over Texas. At first glance, this appears to be cirrus; however, three factors indicate that the streaks are composed of cumulus clouds with separations near the limit of resolution of the camera:

1) The pattern exists only over land. If the clouds were indeed cirrus, we would expect to find no distinct change in pattern along the coastline (Anderson et al., 1966).

2) Although some stations in the area report scattered cirrus, all stations report cumulus or towering cumulus.

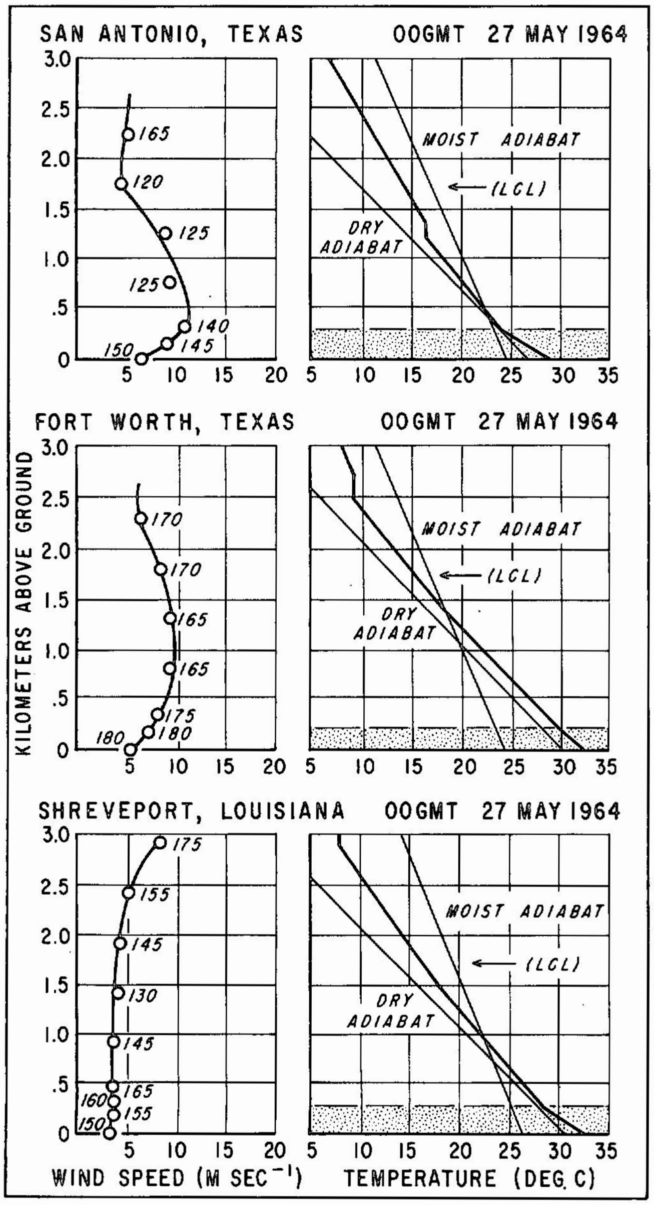


FIGURE 15.—Selected wind and temperature profiles at 00 GMT on May 27, 1964. Heavy line is the actual temperature sounding. Superadiabatic layers are shaded. Numbers give wind direction in degrees. Wind profile for Fort Worth is for 18 GMT on May 26, 1964.

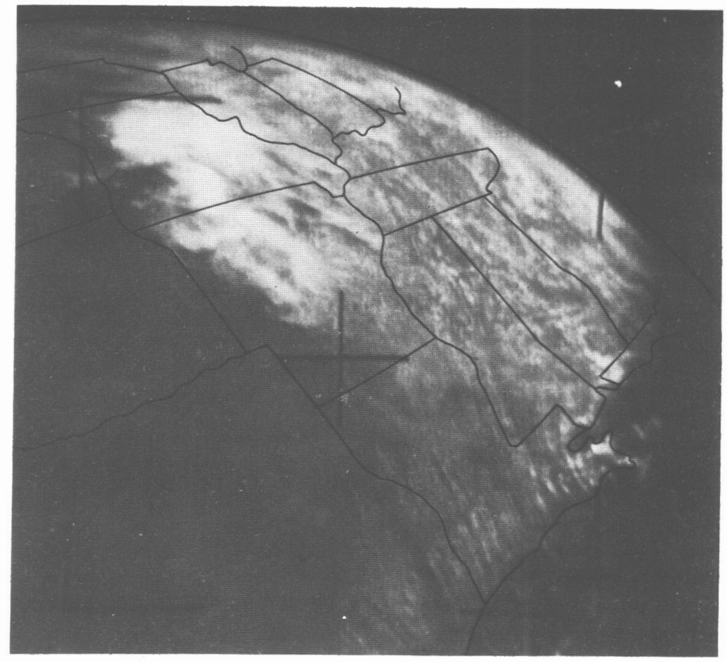

FIGURE 16.—Cumulus bands over eastern Texas and southern Louisiana. TIROS-VII view, 1911 GMT on June 7, 1964.

3) Temperature soundings 4 hr after the picture time indicate cumulus formation at 4,000 to 6,000 ft.

The spacing of the bands in figure 14 is approximately 13 km. Band orientation is parallel to the surface wind or to the wind shear within the convective layer—roughly from 1.5 to 4 km. Brighter clouds in the picture correspond to the regions of shower or thunderstorm activity shown in figure 7. Banded convection in Texas gives way to a more nearly cellular pattern over Louisiana, suggesting a comparison of wind and temperature profiles in the regions of banded and unbanded convection.

Figure 15 shows the soundings at San Antonio and Fort Worth, Tex.—within the region of cloud bands—and, for comparison, the wind and temperature data at Shreveport in northwestern Louisiana. At all three stations, lapse rates are superadiabatic within the first few hundred meters and then nearly dry adiabatic to at least 1 km. The lifting condensation level (LCL) lies between 1.5 and 2 km. The only significant differences are in the vertical variation of the wind within the first few kilometers. At San Antonio and Fort Worth, the wind speed increases by 4 to 7 m sec$^{-1}$ in the first 400 to 800 m above the ground and then decreases by about the same amount above this level. At Shreveport, the wind speed is almost constant with height from the surface to 2 km.

Figure 16 shows distinct cumulus bands over Louisiana and extreme eastern Texas on June 7, 1964. Figure 17 gives wind and temperature profiles at Lake Charles—in southwestern Louisiana—4 hr after picture time. Bands are again parallel to the low-level wind, and band spacing is approximately 15 km. The lapse rate is superadiabatic from the surface to 600 m, and there is a slight increase in wind speed within the first 500 m of about 2 m sec$^{-1}$.

In laboratory experiments (Brunt, 1951), longitudinal rolls develop in *heated flows* with vertical shear. Kuo (1963) has shown analytically that unstable modes of perturbations exist with *negative* stability and uniform wind shear. For large, negative Richardson numbers (marked instability or weak shear), transverse and longitudinal modes are excited, giving rise to patterns similar to the dual mode of convection observed in tropical cumuli by Malkus and Riehl (1964). If the Richardson number is small but negative (strong shear) transverse modes are inhibited. Convection occurs as a series of longitudinal rolls that should lead to cloud streets of the type shown in figures 14 and 15.

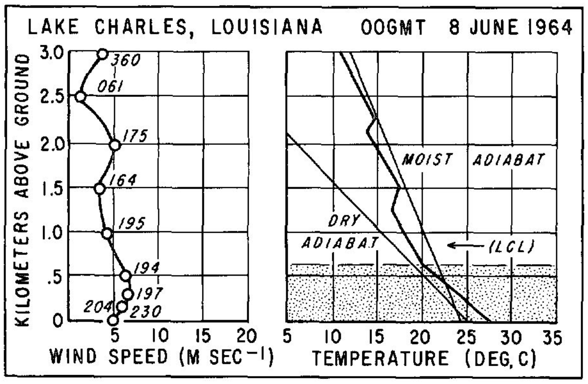


FIGURE 17.—Wind and temperature profiles at Lake Charles, La., at 00 GMT on June 8, 1964. Explanation as in figure 15.

Given the uncertainties introduced by 2- to 4-hr time discrepancies between wind and cloud observations, data in figures 14 through 17 are at least consistent with the hypothesis that banded convection occurs with negative stability and positive vertical wind shear.

Cumulus bands outside the Tropics are usually found in cold outbreaks with a strong positive heat flux from the ground to the air (Kuettner, 1959; Hubert, 1966). In this study, too, cumulus bands occurred only in regions where the air was being heated from below. On both May 26 and June 7, the air was streaming northward from the Gulf of Mexico—warming as it moved over the land surface during the afternoon. Sea-surface temperature in the Gulf at this time of year is approximately 26°C, and air temperatures inland rose on both days to between 30° and 35°C, indicating a warming of the surface air by 4° to 9°C as it moved inland from the Gulf.

## STRATUS BANDS

On several occasions, small-scale bands were apparent in stratus clouds over Texas. Figure 18 is an example of a complicated pattern that developed during dissipation of a solid stratus overcast on May 5 (see also fig. 3). The bands in northwestern Texas resemble mountain waves (Fritz, 1965) but occur well to the east of where such waves could be expected. They appear, instead, to be stratus or stratocumulus in bands perpendicular to the low-level flow. Spacing of the bands is 20 to 25 km—about 25 times the appropriate wavelength for billow clouds (Haurwitz, 1941).

Stratus bands in central Texas parallel the low-level wind with an irregular spacing of approximately 30 km. Figure 19 shows the wind and temperature profiles at Midland, Tex., 2½ hr before the picture time. There is a strong low-level jet at the level of the temperature inversion. The lapse rate is adiabatic within the subcloud layer and superadiabatic within the cloud layer itself.

The presence of strong wind shear and neutral stability at 06 CST suggests that the stratus bands may result from the type of eddying motions which Kuo describes. Surface heating after sunrise may cause the lapse rate to become slightly superadiabatic, and small-scale turbulence may be replaced by convective eddies in the form of longitudinal or transverse rolls. Stratus could be expected to dissipate first in the downwind branches of adjacent rolls, giving a banded appearance to the stratus clouds. As evidence for this, Kaimal and Izumi (1965) show, from Texas tower data, the breakdown of a nocturnal inversion with the subsequent development of a superadiabatic layer within the lowest 300 m after sunrise.

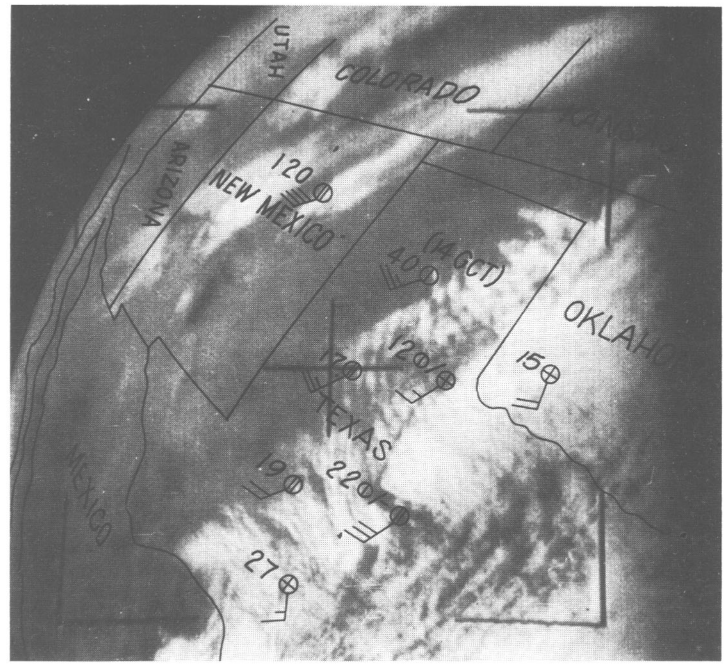


FIGURE 18.—Stratus bands in Texas. TIROS-VII view, 1433 GMT on May 5, 1964.

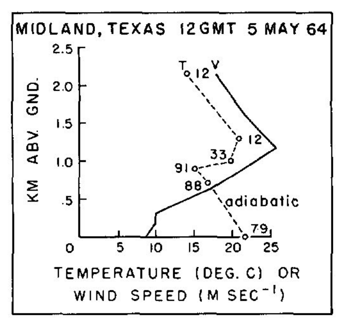


FIGURE 19.—Wind and temperature profiles in the vicinity of the jet, Midland, Tex., 12 GMT on May 5, 1964. Sounding is approximately adiabatic below the jet with a pronounced temperature inversion near the level of maximum wind.

## 5. CONCLUSIONS

The primary conclusions from section 3 have already been stated. We restate only the conclusion that the low-level jet may be responsible for the intrusion of stratus deep into the Central States. The mechanism by which this is accomplished is probably a combination of advection of moisture northward from the Gulf of Mexico and turbulent breakdown of the nocturnal inversion by strong wind shear beneath the core of the jet.

With specially modified radiosonde equipment and with serial ascents at intervals of 1 to 2 hr, Gifford (1952) described the formation of a nocturnal jet and the breakdown of ground-based inversion during a single night at Silver Hill, Md. Between 01 EST and 03 EST, the jet reached a speed of about 40 kt, the temperature rose by 3° at the surface and dropped by the same amount at 1,000 ft. Gifford attributed the temperature changes to turbulence created by the strong boundary-layer wind shear and commented that if the air were sufficiently moist, this condition at 1,000 ft could lead to the formation of stratus clouds.

The same phenomenon is shown in Texas tower data. Izumi (1964) and Gerhardt (1962) show cases with a pronounced cooling at about 300 m and the intensification of an inversion above this level after development of a strong low-level jet. In Izumi's case, the temperature drop at 300 m was 5°C in 1½ hr. Thus, it seems likely that the development of a strong nocturnal jet leads eventually to breakdown of the surface-based inversion, cooling between about 200 and 500 m, and formation of stratus clouds in the moist air moving northward from the Gulf. The critical Richardson number for breakdown of the surface inversion is estimated to be about 0.25 (Lyons, Panofsky, and Wollaston, 1964; Hoecker, 1965).

Data on cumulus cloud bands were highly limited and can only be used as an indication that further satellite studies—especially with ATS pictures—over the New England and Gulf Coast areas might delineate the important parameters for banded convection in the atmosphere. With ATS pictures, time and space resolution would be improved and pictures could be obtained—at least during the summer—near the standard radiosonde observation time of 00 GMT.

Our results suggest, in agreement with Kuo, that necessary conditions for band formation include thermal instability and vertical wind shear. Over land areas there is almost always shear within the boundary layer, and the more critical parameter for band formation may be heating of the air to the point where free convection occurs. When this happens, thermals may originate from near the ground, extending through the region of boundary-layer wind shear. The rising air will then follow a helical path along the flow with clouds forming in the regions of rising motion if and only if the height of the eddy extends above the lifting condensation level.[2]

From this point of view, critical conditions for band formation apply to the subcloud layer and not to the cloud layer itself. The inferences to be drawn from the existence of cloud bands on satellite photographs apply more to the large-scale synoptic process and the lapse rate than to the vertical wind shear.

Wind soundings in the regions of cloud bands did show low-level wind maxima similar to those which Kuettner describes. However, Kuo's analysis of Couette flow shows that curvature of the wind profile is not a *necessary* condition for band formation.

[2] This type of motion has been repeatedly demonstrated in the laboratory (Brunt, 1951). Woodcock (1942) describes a similar motion in the atmosphere with large air-sea temperature differences and strong winds.

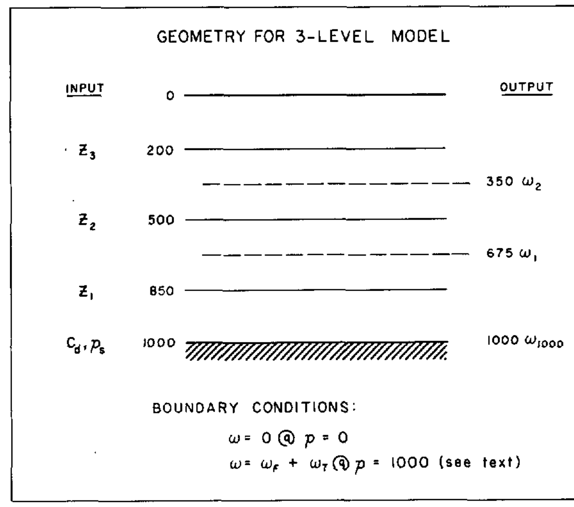


FIGURE 20.—Geometry for three-level model.

The curvature that Kuettner finds may be only a characteristic of the situations in which bands form and not, in itself, a cause of banded convection. Kuettner states that bands in the New England area occur with strong, cold outbreaks, and that a characteristic of such flows is a strong geostrophic jet maximum at the top of the boundary layer. In the South Central States, southerly jet maxima are not associated with strong surface heating, and there was a notable lack of cloud bands in the region of the low-level jet. Bands occurred, instead, in the afternoon just inland from the Gulf where, just as in the Boston area, the air was being heated strongly from below.

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## APPENDIX

### CALCULATION OF VERTICAL VELOCITIES

A three-level, quasi-geostrophic model was used to compute the vertical velocities. Geometry of the model is summarized in figure 20. The $\omega$ equation (Thompson, 1961)

$$\sigma\nabla^2\omega-\frac{f^2}{g}\frac{\partial^2\omega}{\partial p^2}=\frac{g}{f}\nabla^2 J\left(Z,\frac{\partial Z}{\partial p}\right)-\frac{\partial}{\partial p}J(Z,\zeta+f)$$

was solved with standard techniques using a 300-km grid. The stability ($\sigma$) was taken from the summary by Gates (1961) of mean summer and winter stabilities over the United States. As required by the quasi-geostrophic equations, $\sigma$ was considered to be a function of pressure alone.

Since the area of interest included the eastern slopes of the Rocky Mountains, it was essential to incorporate the effects of sloping terrain and of frictional convergence at the lower boundary. We can write

$$\omega_{1000}=\omega_T+\omega_F$$

where subscripts $T$ and $F$ refer to terrain and frictional effects, respectively. The component $\omega_T$ was computed as the scalar product of the 850-mb geostrophic wind and the gradient of standard atmosphere pressures at terrain height,

$$\omega_T=\mathbb{V}_{850}\circ\nabla p_s.$$

Mean terrain heights were taken from McClain (1960). The 850-mb wind was used for simplicity since this is one of the imput levels in the model (fig. 20). However, the 850-mb wind is not a bad approximation to the surface wind since where $\nabla p_s$ is large, $p_s$ is between about 850 and 950 mb. A more serious error lies in assuming that the wind is geostrophic.

The frictional component $\omega_F$ was computed from

$$\omega_F=\rho\frac{g}{f}\left[\frac{\partial}{\partial y}\left(C_d u\sqrt{u^2+v^2}\right)-\frac{\partial}{\partial x}\left(C_d v\sqrt{u^2+v^2}\right)\right]$$

(Cressman, 1960). The surface drag coefficient $C_d$ was specified at each grid point from the highly smoothed values which Cressman presents.

Certainly, both boundary-layer velocities, $\omega_T$ and $\omega_F$, are subject to large errors. However, as Haltiner et al. (1963) have shown, both effects decay quite rapidly with height, so that by about 700 mb the boundary values of $\omega_T$ and $\omega_F$ exert at best a modifying influence on the $\omega$ fields where $\omega$ is large.

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